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Working Paper No. 12

GEOTHERMAL ENERGY

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A Report to the
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By

Karen Barclay
Environmental Engineering Division
The Montana Energy and MHD Research and Development Institute, Inc.

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INTRODUCTION

Geothermal energy, the natural heat of the earth, is a vast potential source of useful power which should become available for a wide variety of important applications as economic exploitation technologies are developed. The decay of radioactive elements deep in the earth's core is the major source of the enormous quantities of heat trapped underground. While the crust of the planet has cooled, temperatures generally increase the deeper one penetrates toward the center. The so-called "normal geothermal gradient" (the average temperature rise as depth increases) is on the order of 1-2^{o}F with each 100 feet of depth--about 75-85^{o}F per mile. Temperatures in the center of the planet may be as high as 8,000^{o}F. The theoretical energy potential of all this heat is staggering (Stone, 1976). In fact, geothermal heat conceivably could be a virtually unlimited source of energy if it were possible to drill deep enough to hit the hot material under every square foot of the earth. But the effort and financial investment necessary for such an endeavor would be enormous. In the foreseeable future, only those few reservoirs close to the surface are likely to be exploited (Herman and Cannon, 1976).

STATE OF THE ART

Today, the exploration and utilization of geothermal resources is receiving a great deal of attention because of the escalating costs for fossil fuels. There are two basic types of geothermal energy: moderate to high temperature sources for electrical generation and moderate to low sources used primarily for space heating. High temperature resources with desirable chemical characteristics for use in steam turbines are rare (even on a world-wide basis). Multiple use development of geothermal resources in the intermediate range (>100^{o}C) probably is limited to electrical energy production and possibly mineral recovery. The world's first geothermal power generating station opened in Laraderello, Italy, in 1940. No other plants existed until 1960 when the Geysers Power Plant in northern California went into operation. Now there are eight countries that together produce 1000 MW of electricity from geothermal heat.

Probably the most important use of moderate to low temperature geothermal heat (<120^{o}C) is space heating. Currently, the potential for utilizing space heating is almost untapped although it has developed rapidly in the past decade. In Iceland, 40 percent of the population lives in houses heated by geothermal water. Japan, New Zealand, the USSR, and the United States are beginning to utilize this resource for heating homes, hotels, apartment complexes, and public buildings.

For instance, geothermal waters at Klamath Falls, Oregon, have provided heat for about 500 buildings, swimming pools, a milk pasteurization plant, and snow melting facilities for roads. The total heat load approximates 5.6 MW. Nearly 400 holes of depths ranging from 27 to 580 meters are

used. Estimated costs for a single well vary from $5,000 to $10,000, and annual operating costs for the wells are less than $100.

In Boise, Idaho, a small geothermal space heating scheme was established in the Warm Springs residential area in 1890. This project received little attention, however, until the 1970's. A joint federal-state effort was initiated early in 1975 to convert existing systems in several state-owned buildings to geothermal water. The aggregate annual fossil fuel expenditure for the ten buildings under consideration for this conversion is $240,000. In addition to a direct economic benefit, the replacement of fossil fuel-based heating with geothermal space heating will affect a significant long-term conservation of fossil fuel. Full implementation of the proposed system would provide year-round heat to the ten state buildings considered for conversion plus thirty additional buildings at less than 70 percent of current fossil-fuel heating costs.

The technology for utilizing geothermal water for space heating is well developed. If the water is high quality (moderate temperature and a small quantity of dissolved solids), it can be piped directly to appropriate building heating systems. When the geothermal water has a high dissolved solids content, heat exchangers must be used to minimize the precipitation and corrosion to which the equipment is subjected.

If the geothermal water temperature is too low for direct utilization schemes, it could be used to preheat water used as input to a hot water heating system. Another method of using this low-grade heat is to use the geothermal water as the heat source for an electric heat pump.

Although Montana has long been known for its geothermal features, actual exploration and utilization of this resource has been limited.

Only one area in Montana has been classified as a known geothermal resource area (Godwin, et al., 1971). The KGRA is located near Yellowstone Park and includes 12,763 acres. An additional 3,834,000 acres shown in Figure 1 have been classified as prospective areas of geothermal resource value.

The majority of the hot springs are located in the southwestern portion of the state where normal faulting is common. The springs are on the low-to-moderate-temperature range and generally have minimal discharges (<1000 gal/min). The indications are that the utilization of the geothermal water probably is limited to non-electrical utilization.

Figure 1

MONTANA GEOTHERMAL RESOURCES AS OF DECEMBER 24, 1970

REFERENCES

McSpadden, William R., 1975, The Marysville Montana Geothermal Project Final Report: Battelle, Pacific Northwest Laboratories.

Facca, G., The Status of World Geothermal Development: U. N. Symposium on the Development and Utilization of Geothermal Resources, Pisa, Proceedings (Geothermics, Special Issue 2) V. 1.

Herman and Connon, 1976, Energy Futures, Geothermal Energy: Inform.

Kunze, J. F., et al., 1976, Geothermal R&D Project for July 1, 1976, to September 30, 1976: EG & G Idaho, Inc. Idaho National Engineering Laboratory, TREE-1030.

Robertson, E. C., Fournier, R. O., Strong, C. P., 1975, Hydrothermal Activity in Southwestern Montana: Second U. N. Symposium on the Development and Use of Geothermal Resources, San Francisco, Proceedings, V. 1.

Balster, C. A., Groff, S. L., Potential Geothermal Resources in Montana.

Barnia, Joseph, 1974, Economics of Multi-Purpose Use of Geothermal Resources: International Conference on Geothermal Energy for Industrial, Agricultural, and Commercial-Residential Uses, Klamath Falls, Oregon, Proceedings.

Sacarto, Douglas M., 1976, State Policies for Geothermal Development, National Conference of State Legislatures, NSF/RA-760230.

Swink, D. G., Schultz, R. J., 1976, Conceptual Study for total Utilization of an Intermediate temperature Geothermal Resource: Aerojet Nuclear Company, ANCR 1260.

Stone, C. D., 1976, The Development of Geothermal Energy: Conference on Energy and the Public Lands, Park City, Utah.

APPENDIX

HYDROTHERMAL ACTIVITY IN SOUTHWESTERN MONTANA

Hydrothermal Activity in Southwestern Montana

EUGENE C. ROBERTSON
U.S. Geological Survey, National Center, Reston, Virginia 22092, USA
ROBERT O. FOURNIER
U.S. Geological Survey, Middlefield Road, Menlo Park, California 94025, USA
CEYLON P. STRONG
New Zealand Geological Survey, Lower Hutt, New Zealand

ABSTRACT

Temperature measurements were made yearly of 14 hot springs from 1966 to 1974; since 1970, measurements have had a precision of 0.01°C. Temperatures range from 45 to 78°C; monthly measurements for 1 1/2 years showed no seasonal fluctuations. Six nearly constant-temperature springs varied only ± 0.4°C from 1970 to 1974; six other springs changed by an average of 1.1°C. Estimated discharges ranged from 0.6 to 60 l/sec; no seasonal effect was found. Chemical compositions of the spring waters remained nearly constant between 1967 and 1974. Silica temperatures obtained from dissolved quartz afford minimum temperature estimates of subterranean water; they range from 59 to 161°C; the changes in silica temperature from 1967 to 1974 ranged from −2.4 to +1.8°C; chalcedony and Na-K-Ca temperatures give similar results. A linear relation was found between surface and quartz temperatures: $T_S = 0.52\ T_Q$; this relation is useful in evaluating the deep hydrology of hot springs.

A 320-km alignment at N 79° E and a 120-km alignment at N 17° E of hot springs are along inferred fault zones; a poorly defined northwest-trending zone of hot springs is roughly along a seismically active zone. Circulation in the breccia of the faults is estimated to reach depths of 3 to 5 km.

INTRODUCTION

The geologic and geophysical environment and the physical and chemical properties of 22 hot springs in and around the Boulder batholith in southwestern Montana were studied to explain their hydrologic and geothermal characteristics. The circulation and heating of the ground water as a reservoir and convection system are related to the geologic structural elements, to the regional and local heat flow from the earth's interior, and to the ancient and present stress distributions which have resulted in permeable fracture channels. Elucidation of these relationships through measurements of temperature, discharge, and chemical composition is necessary to evaluate the hydrothermal systems that produce the hot springs as resources and as adjunct features to local earthquakes.

GEOGRAPHIC AND GEOLOGIC SETTING

The locations of the hot springs are shown in Figure 1 and are listed in Table 1. (Two-letter symbols for springs, given in Table 1 for easy and quick identification, are used throughout this paper.) Two alignments of the springs are obvious—one at N 79° E, 320 km long, and the other at N 17° E, 120 km long. Shorter alignments can also be seen. These alignments are not arbitrary, as there are no other hot springs in the area. A poorly-defined alignment at about N 30° W seems to be related to earthquake activity between Three Forks and Helena.

Inspection of ERTS-1 imagery (NASA, 1972) reveals a lineament of gullies and probable fault scarps between springs BZ and EH with faint extensions to the ends of the alignment at N 79° E. The alignment at N 17° E from springs NB to AL is along the eastern fronts of the Highland Mountains and of the mountains west of Little Whitehead Creek valley, and along Boulder valley. Evidently the hot-spring alignments are controlled by faults. This is not surprising, as Waring, Blankenship, and Bentall (1965) and others have described many such arrays of fault-controlled springs around the world.

The springs in general are in or near the Laramide intrusive bodies, the Boulder batholith and its satellite plutons (Fig. 1), but as shown by a complified diagrammatic geologic section along the BC-HU alignment of Figure 2B, no other geologic correlation can be made, for example, with Cretaceous or Eocene volcanic rocks. (For geologic descriptions, see Klepper, Weeks, and Ruppel, 1957; Becraft, Pinckney, and Rosenblum, 1963; Ruppel, 1963; and Smedes, 1966.) There are no large fault scarps along the BC-HU alignment where it cuts across several mountain ranges (Figure 2A), but small scarps are indicated by the ERTS-1 lineaments, and the alignment may be along an inactive fault. The fault was probably produced by an upper mantle process, judging by the proximity of the Moho (about 40 km depth) relative to the length of the alignment (320 km) and by the fact that a permeable breccia zone to a depth of 5 km is necessary to provide a pathway for the water. Similar statements can be made about the NB-AL hot springs alignment.

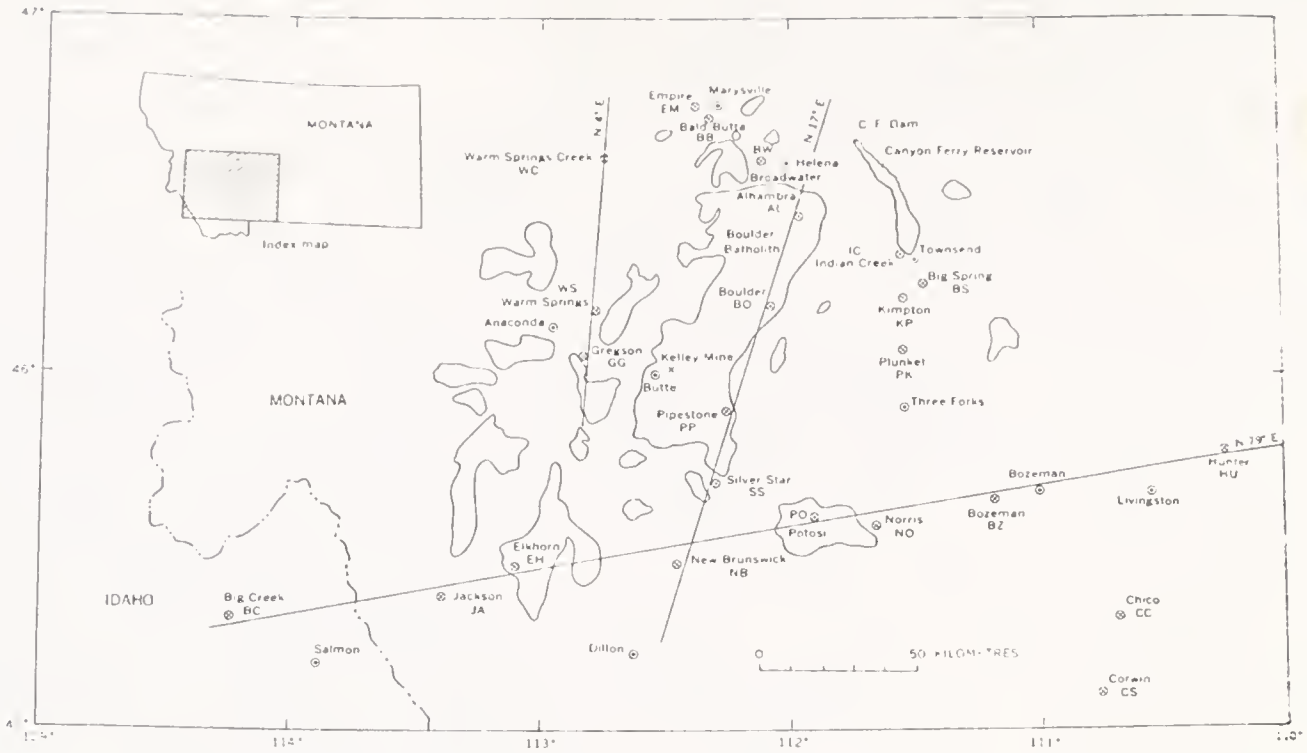

Figure 1. Map of locations and alignments of hot springs in southwestern Montana. The intrusive rocks of the Boulder batholith and associated plutons are outlined. Two-letter symbols for springs are as in Table 1.

Table 1. Hot springs of southwestern Montana.

Name	Symbol	Longitude (West)	Latitude (North)	Water temperature (°C) Surface	Silica Chalcedony	Silica Quartz	Na-K-Ca	Discharge (l/sec)*
Alhambra	AL	111.98	46.45	59	84	114	133	4
Bald Butte†	BB	112.35	46.72	18		—		—
Big Creek	BC	114.26	45.31	93	137	161	175	5
Big Spring	BS	111.48	46.32	15				2000
Boulder	BD	112.10	46.20	76	113	139	125	30
Bozeman	BZ	111.19	45.66	51	83	114	84	8
Broadwater	BW	112.11	46.60	65	107	135	120	1
Chico	CC	110.70	45.32	41	51	85		6
Corwin	CS	110.77	45.10	64		—		1
Elkhorn	EH	113.11	45.46	49	74	105	108	2
Empire†	EM	112.39	46.76	—		—		—
Gregson	GG	112.81	46.04	73	98	127	119	5
Hunter	HU	110.26	45.76	61	82	112	102	30
Indian Creek	IC	111.57	46.35	24		—		3
Jackson	JA	113.40	45.37	59	70	101	145	60
Kelley mine‡	KM	112.54	46.02	50	51	85	83	—
Kimpton	KP	111.56	46.22	18				6
New Biltmore	NB	112.47	45.46	54	65	97	78	2
Norris	NO	111.68	45.58	45	101	129	138	2
Pipestone	PP	112.23	45.90	60	85	116	114	1
Plunket	PK	111.57	46.07	18				300
Potosi	PO	111.90	45.59	51	67	99	63	4
Silver Star	SS	112.29	45.69	71	116	142	120	0.5
Warm Springs	WS	112.79	46.18	78	77	108	81	2
Warm Springs Creek	WC	112.77	46.62	26	26	59	31	1

*Equivalents: 1 gal/min = 63.08 × 10^{-6} m^3/sec; 1 m^3/sec = 10^3 l/sec = 10^3 kg/s of water.
†At 200 m depth in drill holes near Marysville.
‡Seepage at 1.4 km depth, Butte

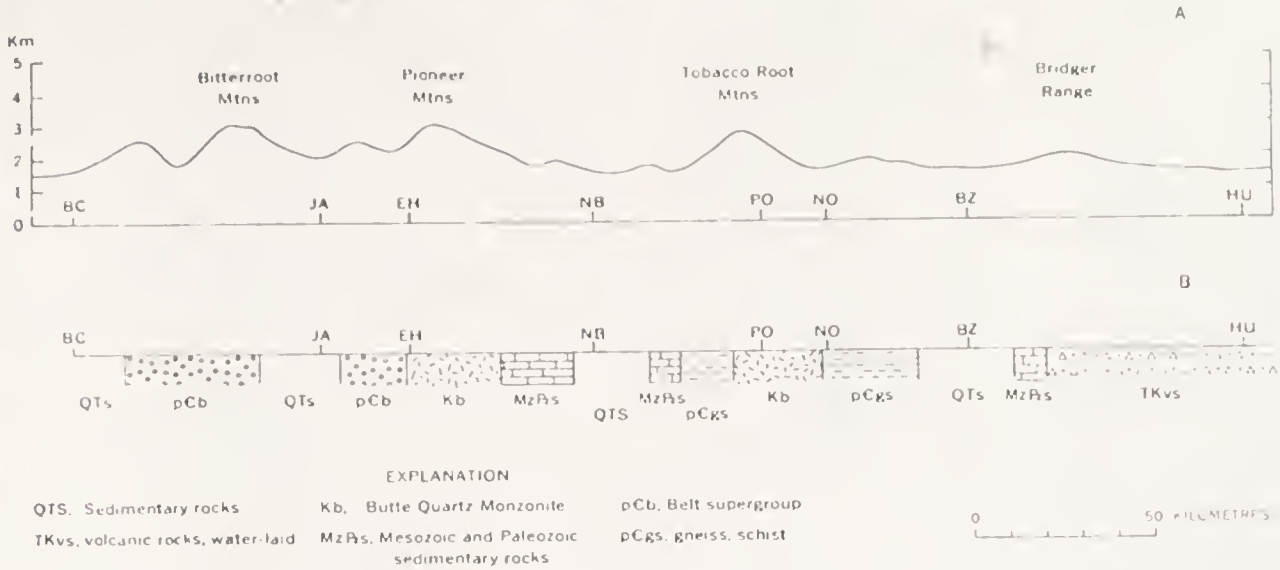

Figure 2. Surface profile (A) and diagrammatic geological profile (B) along the hot-spring alignment at N 79° E. Profile (A) has a vertical exaggeration of 8. Profile (B) is modified from Ross, Andrews, and Witkind (1958).

TEMPERATURE MEASUREMENTS

The heat flow through the crust under the northwestern United States and the Basin and Range province is high—92 MW/m^2 (2.2 cgs units, 10^{-6} cal/cm² sec)—compared with the world average of 61 MW/m^2 (1.5 cgs units). In the vicinity of the Boulder batholith, the average heat flow is 82 MW/m^2 (1.95 cgs units; Blackwell and Robertson, 1973); very high values—816 MW/m^2 (19.5 cgs units) and 370 MW/m^2 (8.9 cgs units)—were measured at EM and at BB (Fig. 1) west of Marysville (Blackwell and Baag, 1973). Deep drilling near EM at Marysville has revealed that the high heat flow there is due to high-temperature circulating water that does not reach the surface (Blackwell, et al., 1974).

Beginning in 1966, and frequently since 1969, temperatures of eleven hot springs (JA, NB, PO, NO, SS, PP, BD, AL, BW, GG, and WS) have been measured with care; five springs (EH, BZ, CS, CC, and WC) have been measured less often; the rest were observed only once. Data on water in mines at Butte were also obtained. Recent temperatures of the water as it issued from the ground at the springs are given in Table 1.

Measurements of the principal study group of the eleven springs listed above were made in 1966 and 1967 with an ordinary mercury thermometer of 2°C precision; a maximum thermometer of 0.5°C precision was used in 1969 and 1970; and a thermistor probe of 0.01°C precision was used from 1970 to 1975. From October 1969 through February 1972, readings were taken once a month to ascertain seasonal effects (if any) resulting from hot-spring water mixing with surface water; no seasonal effect was found in any of the springs.

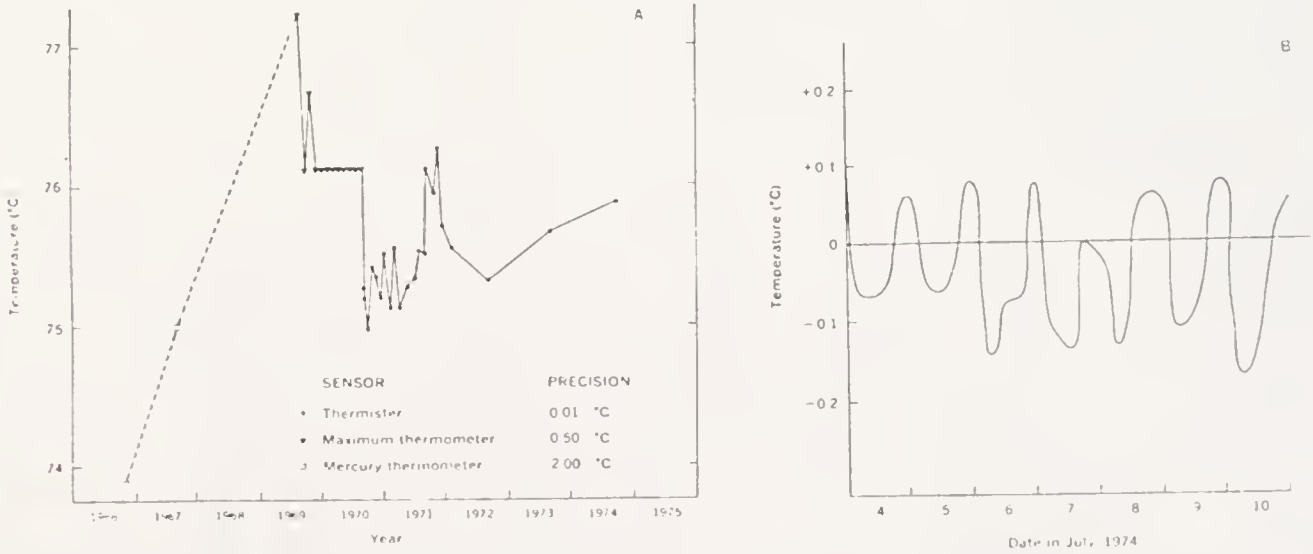

Figure 3. Water temperatures of Boulder hot spring. *A*: Chronological plot, showing the temperature sensors used and their precision. *B*: Diurnal variation during one week in July 1974, due to drawdown for resort use.

Temperature data for spring BD of the Ranch[illegible] resort at Boulder are plotted in Figures 3A and 3B. There seems to be a 1 to 2 year cycle of temperature between 75 and 76°C in BD water (Fig. 3A). To study more closely the variation with time at BD, continuous recording has been made of temperature to 0.01°C intermittently from 1969 to 1973, and nearly completely from 1973 to the present; a thermistor probe and a standard resistor are used alternately in a balanced bridge circuit. A normal daily fluctuation of 0.1°C due to water usage by the resort's customers is shown in Figure 3B. The wider fluctuations, to 0.4°C in a month's time (Fig. 3A), are thought to be due to changes in circulation and to mixing of water from different channels in underground fractures.

The temperature record of spring SS in Figure 4 demonstrates a decrease, with some fluctuation, from 73 to 71°C in the 5-year period, 1969 to 1974. The withdrawal of water by the resort owners from both springs, BD and SS, was restricted to the natural flow, so temperature changes greater

Figure 4. Chronological plot of water temperature at [illegible] Star hot spring. Symbols have same meanings as in Figure 3A.

Figure 5. Chronological plot of water temperature at Boze[illegible]man hot spring. Symbols have same meanings as in Figure 3A.

probably reflect the natural changes in permeability of closing and opening of channels by displacements of rock, breccia, and gouge along faults.

In contrast the steady decrease in temperature from 59 to 47°C (Fig. 5) of spring BZ over the same 5-year period is due to interference with the natural system. Two holes were drilled to 20 m and one to 120 m, and pumps were installed; withdrawal of hot water was increased from the natural discharge of 3.2 l/sec (50 gal/min) to 7.6 l/sec (120 gal/min).[1] Apparently, time for heat transfer from rock to water has been reduced, and a drop in water temperature has resulted. Interference by drilling of holes disrupted in a different way the natural system at Brady's Hot Springs, Nevada, the normal flow of 1.3 l/sec (20 gal/min) was reduced to nothing at the spring, many small steam vents were created, and the water was dispersed widely (F. H. Olmsted, oral commun., 1975).

TEMPERATURES AND DISCHARGES

Samples of water were taken for chemical analysis from many of the hot springs both in 1967 and in 1974. The latter set of analyses is listed in Table 2. The chemical composition remained very nearly the same during the period. In general, these springs have low salinity. Spring NB has the highest content of dissolved solids, but the salinity is <0.2% for all springs.

A useful method of ascertaining the temperature at depth of circulating water is to use the solubility of silica as a thermometer (Fournier and Rowe, 1966); another method utilizes ratios of dissolved Na-K-Ca (Fournier and Truesdell, 1970). These temperatures are listed in Table 1. Temperatures of last equilibration of the water in dissolution of quartz (or by a separate relationship, dissolution of chalcedony) are minimum estimates of the water temperature at depth.

The water temperatures derived from the solubility of quartz, and surface temperatures measured at the spring mouths are plotted in Figures 6 and 7 with distance along the two northeast alignments. The figures show, on a small scale, the changes in the quartz and the surface temperatures from 1967 to 1974. The quartz-temperature changes over the seven years ranged from −2.4 to +1.8°C and averaged +0.11°C ± 1.78°C (standard deviation) for the 12 springs so measured. Such changes are barely noticeable in Figures 6 and 7.

Assuming the normal temperature gradient in the Boulder batholith area to be 30°C/km (Blackwell and Robertson, 1973), circulation of the water to depths of 3 to 5 km would heat the water to the quartz temperatures found. Greater depths of circulation cannot be repudiated, as the silica-solubility method affords only a minimum value. The four alignments of the springs imply that circulation is in the breccia of deeply penetrating faults.

The changes in the surface temperatures of the springs were not large (Figs. 6 and 7), some temperatures remained relatively constant and others drifted. We shall examine the precise measurements (0.01°C), made from 1970 to 1974; measurements were made at monthly intervals in 1970–1971. The mean surface temperatures and quartz temperature changes of the six springs having nearly constant surface temperatures are listed in Table 3, the standard deviations

[1]One l/sec = 15.9 gal/min.

Table 2. Chemical analyses of hot spring waters (10^{-3} kg/m^3 = 1 ppm).

Name	SiO_2	Ca	Mg	Na	K	Li	HCO_3	CO_3	SO_4	Cl	F	B
Alhambra	64.7	18 8	3 2	215	10 0	0.38	505	9.0	93 0	8.5	4 8	0.2
Big Creek	150.0	5.3	0 2	220	14.0		488	0	53 0	29.0	15 0	
Boulder	104.0	2.3	0.01	124	3.8	0.26	198	2.7	77.0	24.0	6.9	0.55
Bozeman	64 0	8 5	1 75	133	3 2	0.04	158		119 0	52.0	7 5	0.15
Broadwater	97.1	11 5	0.75	188	6.0	0.55	298	3.6	189 0	21.5	6 2	0.82
Elkhorn	53 6	2 0	0	49	1 0	0 05	99	2 1	32.0	8.5	1 6	
Gregson	81.2	4 0	0.04	181	4 1	0.70	188	3 1	195 0	10.0	11.0	0 3
Hunter	62 2	0.6	0 03	88	1.0	0.04	196		18 5	24 0	4 6	0.7
Jackson	49 3	10.5	3.2	248	11.0	0.35	632	20.6	50.0	11.0	1.3	0.65
New Biltmore	45.2	280.	79.	163	26.0	0.21	258		1180 0	52.0	1 8	
Norris	86 8	19.	2.6	208	10.6	0 10	397	9 4	147.0	27.0	4.4	0.20
Pleystone	66.4	2.6	0	101	2.0	0.09	134	0 4	99.0	27.0	3.1	0 50
Potosi	47 0	10.6	0.01	94	1.9	0.06	84	1.7	155.0	4.4	3.6	0
Silver Star	109	9 6	0 52	173	6.9	0.37	185	0.2	193.0	15.0	5.4	0.45
Warm Spring	56.4	190.0	22.	131	23.2	0.42	216		658.0	39.5	2.4	0.30
Warm Springs Creek	17.9	130.0	38.4	23	5.6	0 16	243		371 0	2.6	0.8	0

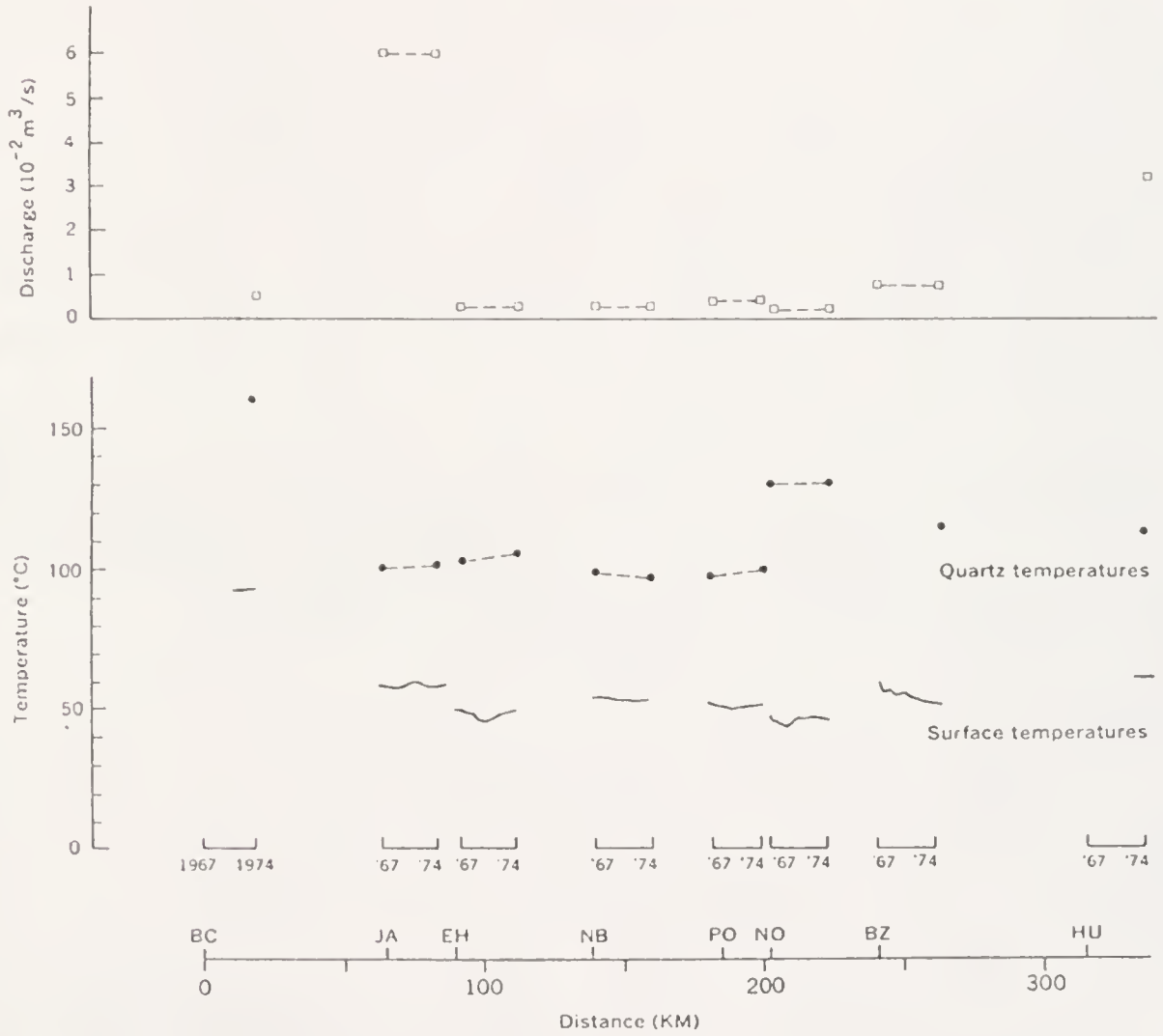

Figure 6. Temperature and discharge data for hot springs along the alignment at N 79° E. The measurements from 1967 to 1974 are shown for each spring.

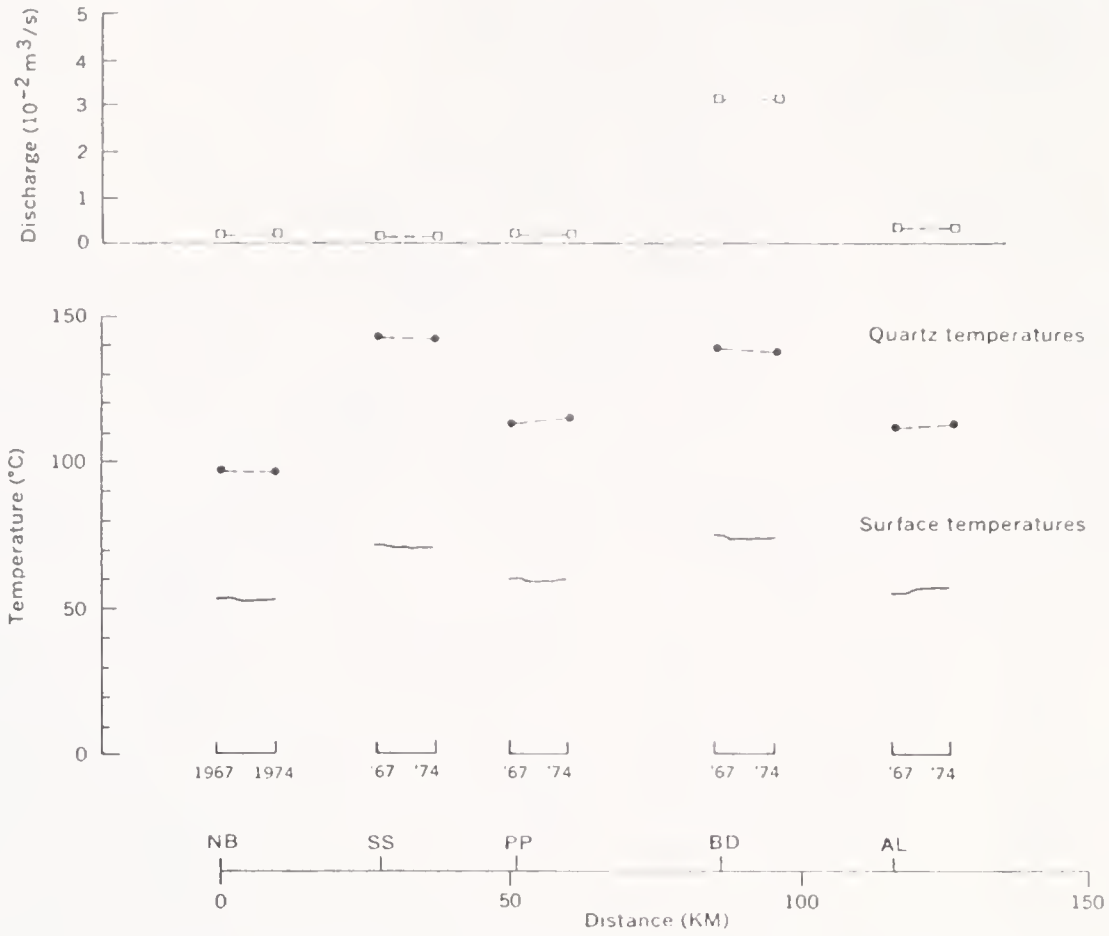

Figure 7. Temperature and discharge data for hot springs along the alignment at N 17° E. The measurements from 1967 to 1974 are shown for each spring.

for the 4-year period, with one exception, are <0.5°C. The changes of surface and quartz temperatures of the seven springs showing temperature drifts are listed in Table 4; aside from spring BZ, which was disturbed, the temperature shifts are <3°C during the 4-year period.

The possibility of finding some spatial or temporal uniformity of temperature or of discharge was the motive for plotting Figures 6 and 7. Apparently there is no significant relation among the observations aside from the relative constancy of the individual spring temperatures and discharges. Waring, Blankenship, and Bentall (1965) state that uniformity in temperature, flow, and composition are very commonly observed, and they cite a statement that the springs near Bône, in Algeria, have not changed more than 4°C in 2000 years.

Although no correlation of temperatures in time and in distance has been found along the alignments, a marked linearity of surface temperature T_S with quartz temperature T_Q is shown in Figure 8; the line projects through zero and has the equation

$$T_S = 0.52\, T_Q \qquad (1)$$

Table 3. Mean temperatures of surface water and quartz temperature changes of hot springs having nearly constant temperatures.

Spring Symbol	Temperature °C Mean	Std. dev.	Quartz temperature change
BW	64.79	0.43	−2.4°C
BD	75.52	0.33	−0.3
PP	60.28	0.21	+1.1
NB	53.49	0.21	−1.0
JA	58.54	0.85	+0.8
WS	78.14	0.47	+1.2
Average values		0.42°C	−0.10°C

Table 4. Surface-water temperatures and quartz and surface temperature changes of certain hot springs that show temperature drift.

Spring Symbol	Temperature °C Aug. 1970	Aug. 1974	Temperature change °C Surface	Quartz
AL	57.89	58.79	+0.9	+0.4
SS	72.01	70.78	−1.23	−1.0
EH	48.08	49.44	+1.36	+1.7
PO	50.00	50.74	+0.74	+1.8
NO	42.68	45.25	+2.57	−0.1
BZ	56.36	51.08	−5.28	n.a.
GG	70.19	72.68	+2.49	−0.9
Average changes (without BZ)			+1.14°C	−0.32°C

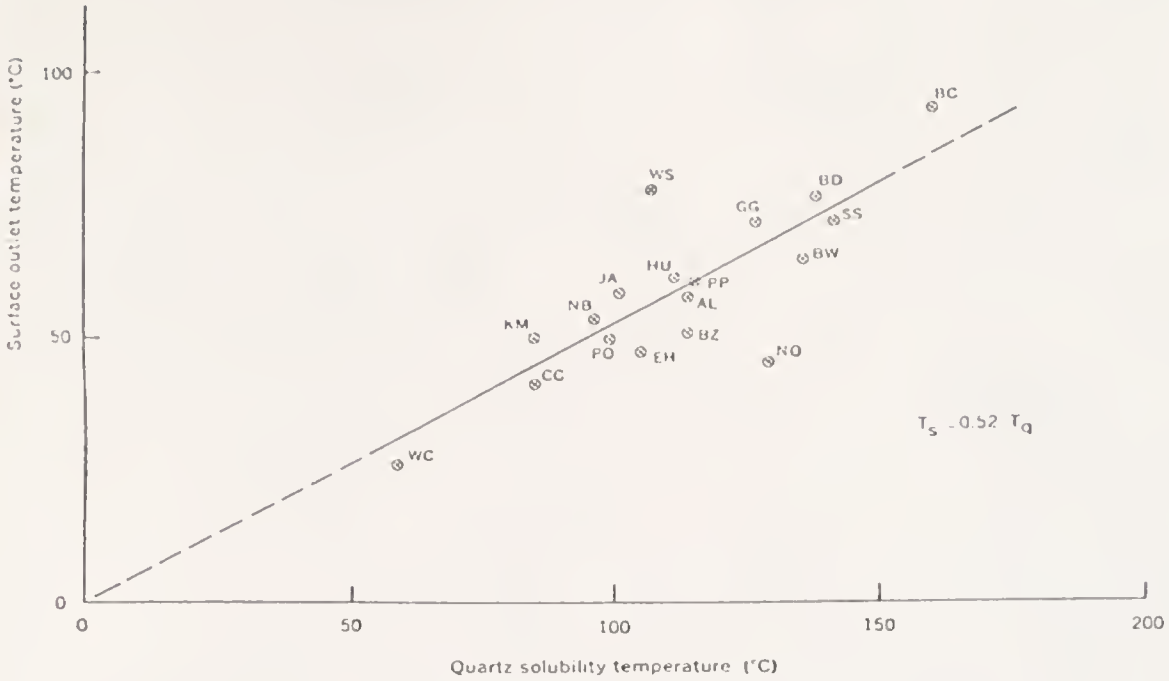

Figure 8. Surface-water temperatures of hot springs plotted against silica temperatures determined by the quartz-solubility method.

(We have found this same line to be true for temperatures of springs in Idaho and Nevada.) The point for spring NO has a low T_S because of evaporative cooling of the large open tank above the egress.

In Figure 8 the point for spring WS is aberrant. We suggest that this is because the water equilibrated at depth with chalcedony rather than with quartz and because the water moved to the surface with little loss of heat on the way. The situation may be similar to that at the travertine-depositing springs at Mammoth in Yellowstone National Park, where water apparently flows to the surface relatively quickly through smooth carbonate-lined channels. Measurements in a hole drilled through one of the huge travertine mounds at Mammoth (White, et al., 1975) reveal a constant temperature of 73°C from depths of 15 m to 100 m. At Mammoth, the dissolved silica in the water appears to be in equilibrium with chalcedony at about 73–75°C. The water discharged from spring WS flowed through limestone beds at depth and has deposited a large travertine cone at the spring outlet, which we estimate required about 1000 years to accumulate. As at Mammoth, the silica in the water of spring WS gives a chalcedony temperature (Fournier, 1973) of 77°C which is in excellent agreement with the measured temperature of 78°C and with the Na-K-Ca temperature of 81°C.

EARTHQUAKE ACTIVITY

Three major earthquakes, one in 1925 (M = 6.75) and two in 1935 (M = 6.0 and 6.25) occurred near Helena, Montana. Their epicenters (Friedline and Smith, 1974) are plotted in Figure 9; in addition, epicenters for three recent nearby earthquakes (M > 4) are shown. The northeasterly hot-spring locations (Fig. 9) show a rudimentary alignment parallel to the northwest trend of the earthquake epicenters; also, this trend parallels that of a system of major Tertiary faults that bound the Bridger Range on the west, the Clarkston Basin on the east, the Townsend valley on the east and west, and the Helena valley on the northeast (R. G. Schmidt, written commun., 1975). The juxtaposition of hot springs and epicenters may therefore signify a common fault system. The temperatures recorded at spring BD become disturbed after local earthquakes, which sometimes temporarily offset the otherwise uniform temperature (Fig. 3B) by > 2°C; the shift is determined by the magnitude and proximity of the shocks. So far, no reliable premonitory temperature changes have been observed that might help in predicting local earthquakes.

A dam was completed in 1954 at Canyon Ferry, and the Missouri River was ponded behind it for 40 km to the southeast. This reservoir is near the northwest-trending earthquake and hot-spring alignment, and although fracture permeability appears to be low in the rocks of the Belt Supergroup cropping out along the banks, water might eventually reach the northwest-trending fault zone and trigger earthquakes. In view of the historical consequences of earthquakes occurring near such reservoirs, we suggest careful monitoring of local seismicity; occasional, low-magnitude (M 3–4) earthquakes are occurring near Townsend and Three Forks.

RESOURCE EVALUATION

The hot springs of southwestern Montana are all of the warm water type, without steam. Their discharges (Table 1) are quite low, < 60 l/sec (< 1000 gal/min), except for two large low-temperature springs, BS and PK, whose waters are used for irrigation. Some of the warm springs may result from the mixing of deep hot water and shallow cold water. If the resulting mixture does not have time to equilibrate chemically before moving up to the surface, the chemical geothermometers may be in error. Too little information is available to try mixing models (Fournier and Truesdell, 1970) to estimate the hot water temperatures. The silica (quartz) temperatures (Table 1) range from about 85 to 160°C.

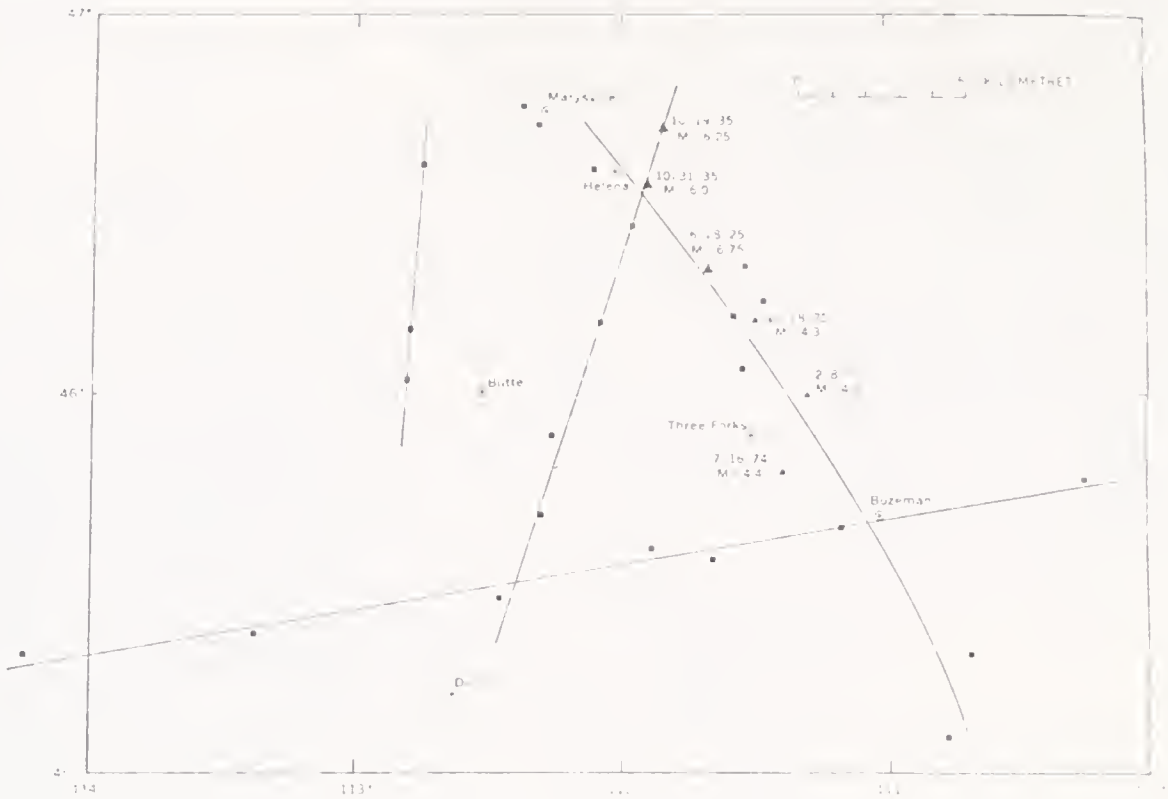

Figure 9. Locations of hot springs and of three magnitude-6 and three recent magnitude-4 earthquakes in southwestern Montana. Symbols: triangle = earthquake epicenter; square = hot-spring location. The alignment from Bozeman to Marysville shows rough parallelism of hot springs and earthquake epicenters; other alignments are as on Figure 1.

the inferred temperatures at depth of the circulating water; these are minimum values, as the temperature may be higher if mixing, boiling, or some other nonadiabatic process occurs. Assuming that no anomalous heat sources are present, at a gradient of 30°C /km the highest temperature water would only be available at depths of 3 to 5 km.

These results indicate that the springs will probably be commercially exploited only for present recreational uses, for heating of buildings, and for low-temperature small-scale industrial purposes. Caution in exploitation is urged, as excessive drilling into the natural systems could permanently disrupt the water flow. Recharge of pumping of water back into the ground should not be undertaken without prior careful study to evaluate the risk of stimulating earthquakes, such as those that occurred at the Rocky Mountain Arsenal well near Denver.

CONCLUSIONS

We have made no attempt to ascertain relationships of the hot springs in southwestern Montana to the water table and to ground water flow. The rainfall is low, about 35 cm/yr (14 in./yr), but the recharge from the large drainage areas of high relief near these springs would suffice for the low discharges of the hot springs. It is our contention that the primary controls of the hydrologic balance of these springs are deep faults responsible for the alignments of springs.

The proximity of the hot springs to the Boulder batholith and nearby stocks may be explained as being the most recent activity in a region in which plutonic injections began 80 million years ago, volcanic eruptions took place 20 million years ago, and mountain building has occurred into the Holocene.

The driving mechanism for the hot springs is probably a combination of the head of cold, descending (meteoric) water and the increase in buoyancy of the deeply circulating water resulting from its thermal expansion and lowered viscosity on heating. Given the high regional heat flow, the high geothermal gradient, the probable high temperature of the upper mantle, the descent of the meteoric waters to depths of 3 to 5 km to attain the observed heat seems reasonable, and no special magma body in the crust seems needed.

We infer from the results of this 7-year study and the experience of others that the springs are in a steady-state condition. In the 11 closely studied springs, we found no seasonal fluctuations in temperature or discharge. From 1970 to 1974, six nearly constant springs varied by ±0.42°C, the average of the standard deviations from the 4-year means of their surface-water temperatures; quartz temperature changes averaged −0.10°C from 1967 to 1974. Seven other springs changed in the same periods by averages of −1.3 and +0.32 C for their surface-water and quartz temperatures.

The implications from the linear relation between surface-water and silica temperatures of Equation (1) are significant to the hydrology of deeply circulating spring waters. First: the silica temperature can be approximately predicted from the surface-outlet temperature.

Second, considering that the discharges of the springs vary by a factor of 100, (from 0.6 to 60 l/sec [10 to 1000 gal/min]), for Equation (1) to be valid, the loss of heat

to the walls must be essentially independent of the volume of flow and the wall-rock type (excepting limestone), heat transfer from the ascending water to the walls must be nearly constant, which implies that a well-adjusted balance exists between surface area of contact and velocity of flow. (The average thermal conductivity of most massive rocks is 3.0 ± 0.5 W/mK.) The obvious limiting factor is that of permeability; high-carbonate waters build smooth low-resistance channels, and heat loss to the walls is greatly reduced.

Finally, as the presence of near-surface cooling magma bodies can be discounted, the maximum temperature attained by the circulating water (which is nominally equal to the silica temperature) is a direct index of the maximum depth of circulation; the local geothermal gradient can be used to calculate depth from the surface water temperature. In general, we suggest that from the fault structural control, the probable drainage recharge area, the water table and lateral surface rock permeability, and from the physical and chemical characteristics of the springs, the hydrology of the hot springs can be reasonably well elucidated.

ACKNOWLEDGMENTS

We are grateful to the owners and operators of the hot springs for their permission and hospitable cooperation with us in our studies: W. L. Bompart, Norman Rogers, J. C. Alley, H. E. McMurtrey, J. Stanovitch, William Harvey, Warren Henschel, Robert McNeill, John Dooling, M. Zankourky, Richard Josephson, and Charles Page. In particular, we are grateful to Mr. and Mrs. J. W. Sanddal, Diamond-S Ranchotel, Boulder, Montana, and Mr. H. E. Wilson, present manager there, for permission to set up continuous temperature-recording equipment at their hot spring. We thank J. M. Thompson for chemically analyzing the water samples for us. We appreciate very much the effort of R. W. Werre and Rudolph Raspet in designing and building temperature-sensing and recording equipment and in calibrating thermistors. Suggestions by M. R. Klepper, G. D. Robinson, R. G. Schmidt, and A. H. Truesdell were very helpful. The encouragement of Jerry Eaton and of M. F. Kane is also gratefully acknowledged. We appreciate very much the effort by James L. Fenton in making monthly temperature measurements which demonstrated that no seasonal effect occurs.

REFERENCES CITED

becraft, G. E., Pinekney, R. M., and Rosenblum, S., 1963, Geology and mineral deposits of the Jefferson City quadrangle, Montana: U.S. Geol. Survey Prof. Paper 428, 101 p.

Blackwell, D. D., and Baag, C. G., 1973, Heat flow in a "blind" geothermal area near Marysville, Montana: Geophysics, v. 38, p. 941–956.

Blackwell, D. D., Brott, C. A., Goforth, T. T., Holdaway, M. J., Morgan, P., Petefish, D., Rape, T., Steele, J. L., Spafford, R. E., and Waibel, A. F., 1974, A brief description of geological and geophysical exploration of the Marysville geothermal area: Natl. Science Foundation Conf. Res. Dev. Geothermal Energy Resources, Pasadena, Proc., p. 98–110.

Blackwell, D. D., and Robertson, F. C., 1973, Thermal studies of the Boulder batholith and vicinity, Montana, *in* Guidebook, Butte field meeting: Soc. Econ. Geol., D1–D8.

Fournier, R. O., 1973, Silica in thermal waters: Laboratory and field investigations: International Symposium on Hydrogeochemistry and Biogeochemistry, Japan, 1970, Proceedings, v. 1, Hydrogeochemistry, Washington, D.C., J. W. Clark ed., p. 122–139.

Fournier, R. O., and Rowe, J. J., 1966, Estimation of underground temperatures from the silica content of water from hot and wet-steam wells: Amer. Jour. Sci., v. 264, p. 685–697.

Fournier, R. O., and Truesdell, A. H., 1970, Chemical indicators of subsurface temperature applied to hot waters of Yellowstone National Park, Wyoming, U.S.A.: UN Symposium on the Development and Utilization of Geothermal Resources, Pisa, Proceedings (Geothermics, Spec. Iss. 2), v. 2, pt. 1, p. 529–535.

Friedline, R., and Smith, R. L., 1974, Contemporary seismicity in the Helena, Montana, region, *in* The Marysville, Montana, geothermal project tech. rept: NSF-RANN Grant No. GI-38972, p. 75–94.

Klepper, M. R., Weeks, R. A., Ruppel, E. T., 1957, Geology of the southern Elkhorn Mountains, Montana: U.S. Geol. Survey Prof. Paper 292, 82 p.

NASA Goddard Space Flight Center, 1972, ERTS-1 imagery, 0.8–1.1 μm band: U.S. Soil Conservation Service, Mosaic Sheet B, scale 1:1 000 000.

Ross, C. P., Andrews, D. A., and Witkind, I. J., 1958, Geologic map of Montana: U.S. Geol. Survey and Montana Bur. Mines and Geology.

Ruppel, E. T., 1963, Geology of the Basin quadrangle, Montana: U.S. Geol. Survey Bull., v. 1151, 121 p.

Smedes, H. W., 1966, Geology and igneous petrology of the northern Elkhorn Mountains, Montana: U.S. Geol. Survey Prof. Paper 510, 116 p.

Waring, G. A., Blankenship, R. R., and Bentall, R., 1965, Thermal springs of the United States and other countries of the world—a summary, U.S. Geol. Survey Prof. Paper 492, 383 p.

White, D. E., Fournier, R. O., Muffler, L. J. P., and Truesdell, A. H., 1975, Physical results of research drilling in thermal areas of Yellowstone National Park, Wyoming: U.S. Geol. Survey Prof. Paper 892, 70 p.